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PERSISTENT LARGE-SCALE FLOW ANOMALIES.

PART I: CHARACTERISTICS OF DEVELOPMENTS

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1. INTRODUCTION

Our objective in these lectures will be to describe and interpret some of the observed characteristics of major large-scale flow anomalies that persist beyond the durations usually associated with synoptic-scale variability. The main focus of Part I will be on the development of the persistent anomalies, while Part II will concentrate on changes in synoptic-scale eddy activity and cyclogenesis that accompany the large-scale flow anomalies.

The basic question considered in Part I is that of the sources for the events, that is: *how do the persistent large-scale disturbances originate* ? We will address this first as an observational problem, and then examine the implications of the results for theories of persistent anomaly development.

For brevity, part I will focus on the development of persistent negative height anomalies that occur during wintertime over the central North Pacific (the PAC negative cases); however, descriptions of additional aspects of the life cycles of persistent anomalies, including analyses of the case breakdowns and of cases in other regions, can be found elsewhere (Dole, 1986b; Dole, 1987). At upper levels, the PAC negative cases are typically manifested synoptically by the formation of an abnormally strong jet across much of the North Pacific, a highly amplified ridge near the west coast of North America and a deeper than normal trough over eastern North America (Dole, 1986a). They are also often related to unusually intense realizations of the Pacific - North American (PNA) teleconnection pattern

(Wallace and Gutzler, 1981). At the surface, the cases are generally characterized by an anomalously intense and large-scale cyclonic circulation that frequently extends across most of the North Pacific Ocean at mid-latitudes (Dole, 1986a, 1986b).

2. SOURCES FOR LOW-FREQUENCY VARIABILITY

Before proceeding to the observational analyses, it is worthwhile to consider some of the potential mechanisms that have already been proposed for generating low-frequency variability. Perhaps unfortunately, there are many possibilities, including:

1) *anomalous thermal or orographic forcing* (e.g., Hoskins and Karoly, 1981) *or mean flow variations* (e.g., Held, 1983; DaSilva and Lindzen, 1987), both of which may generate transient low-frequency Rossby waves.

2) *large-scale flow instabilities.* As we will see, at large scales instabilities of zonally symmetric mean flows are unlikely to achieve sufficiently rapid growth rates to account for the observed developments. However, studies that include zonal asymmetries in the basic state, such as of the barotropic instability (Simmons *et al.*, 1983) or baroclinic instability (Frederiksen, 1983) of the wintertime mean flow, have been able to identify unstable modes that have growth rates and structures in closer agreement with observations.

3) *non-linear interactions with synoptic-scale disturbances.* This mechanism has received particular attention in studies on blocking. As will be discussed in Part II, most early investigations (e.g., Green, 1977; Austin, 1980) focused on the potential role of synoptic-scale disturbances in the *maintenance* of blocking. More recently, however, considerable evidence has been presented suggesting that at times large-scale flow anomalies may develop or be intensified through systematic interactions with synoptic - scale disturbances. Although most research in this area has focused on essentially barotropic

effects of eddy momentum (or vorticity) fluxes on pre-existing large scale flow anomalies (Shutts, 1983; Hoskins *et al.*, 1983; Illari, 1984; Haines and Marshall, 1987; Holopainen and Fortelius, 1987; Mullen, 1987), some general circulation model (e.g., Gall *et al.*, 1979; MacVean, 1985; Young and Villere, 1985) and observational studies (Hansen and Chen, 1982; Hansen and Sutera, 1984; Colucci, 1985, 1987), have presented results suggesting that at times baroclinic effects may also play an important role. In addition, theoretical studies suggest that even effects (due either to physical processes or to transient eddies) that formally appear dissipative may at times result in amplification of waves having scales and structures comparable to observed persistent anomalies (e.g., Held *et al.*, 1986). In Part II, we will return to this last point when interpretations of time-mean budget results are considered.

Although our current list of mechanisms for low-frequency variability is already relatively long, it is by no means exhaustive, and other mechanisms can readily be envisaged. It seems plausible, however, that the potential for skill in long-range prediction will depend significantly on the relative importance of the various mechanisms for producing low-frequency fluctuations (as well as on our ability to correctly simulate them), and therefore, that the identification of the major processes contributing to the genesis of persistent flow anomalies presents an important problem for both observational and theoretical meteorologists.

In reviewing the recent literature (particularly that related to the development of the PNA pattern), it is my impression that present interpretations of the developments of mid-latitude persistent anomalies have been derived mainly from barotropic or equivalent barotropic models (perhaps the most commonly mentioned mechanisms being either barotropic instability of the time mean flow or essentially barotropic Rossby wave dispersion from persistent anomalous sources, with variations in tropical heating being perhaps the most frequently cited source). Considerable evidence has been presented in support of these interpretations (see, e.g., Wallace and Lau, 1986, for an extensive review of the role of barotropic instability in low-frequency fluctuations, and Weickmann *et al.*, 1985, for an indication of the possible importance of tropical diabatic heating on extratropical variability in the 30-50 day time scale range), and it now appears likely that these processes play a significant role in the generation and maintenance of some of the low-frequency disturbances observed in the extratropical Northern Hemisphere wintertime circulation.

My emphasis, here, however, will be mainly on points of *discrepancy* between observations and current theories, focusing particularly on the adequacy of barotropic models in accounting for major characteristics of the developments. I will also describe some additional processes that have so far not been considered but that I speculate may be important in the initiation of certain major cases of persistent flow anomalies. In particular, on the basis of observations I will suggest that:

- much of the temporal variability associated with the major persistent anomaly patterns (such as the PNA pattern) is realized during energetic but relatively transient events;

- in many cases, baroclinic processes appear to play an important role throughout the developments, while barotropic processes are likely to provide significant additional contributions at a later stage of the developments and in the maintenance of the anomalies, once they are established;

- variations in the response to tropical heating (which may be due to either heating or mean flow changes) may often be initiated by events that have mid-latitude origins.

In concert with a basic theme of this seminar, I will also present evidence suggesting that at times synoptic-scale disturbances play an important role in triggering or catalyzing the larger-scale flow changes.

3. DATA AND METHODS

Both the data base and analysis methods have been discussed extensively elsewhere (Dole and Gordon, 1983, hereafter DG; Dole, 1986a, hereafter DS) so only the main points will be summarized here. The basic data are obtained from the twice-daily (0000 GMT and 1200 GMT) National Meteorological Center (NMC) final analyses of the Northern Hemisphere sea-level pressure as well as geopotential heights, winds and temperatures at all standard levels from 1000 mb to 100 mb for the 14 120-day winter seasons (beginning 15 November) from 1963-1964 through 1976-1977.

As in the previous studies, cases were selected by applying very simple, general objective selection criteria (Fig. 1) to determine all events when anomalies of a given sign persisted beyond some threshold value M for at least a given number of days T.

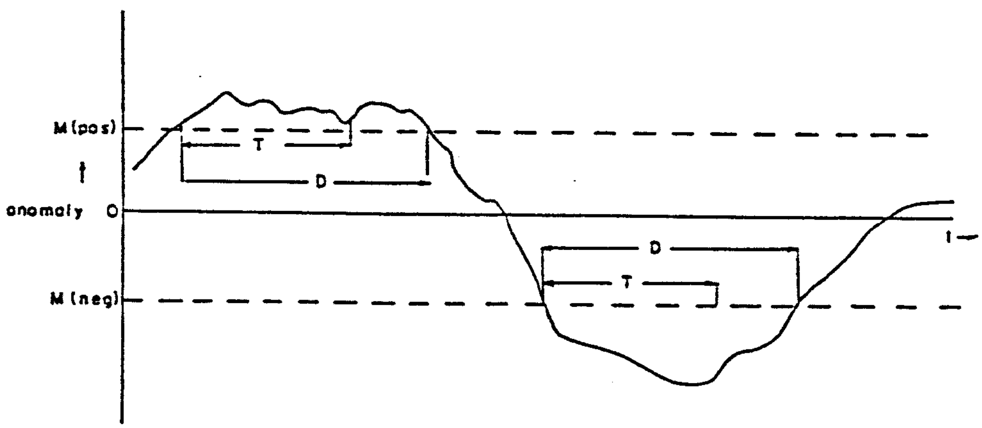


Fig. 1. Illustration of method used for defining cases. A persistent positive (negative) anomaly case of duration D satisfying criteria (M ,T) is defined at a given point if the anomaly at that point exceeds (is less than) the threshold value M for at least T days. Examples are given for both positive and negative anomaly cases.

As in DG, anomalies are defined as departures from seasonal trend values. For the 500 mb height analyses alone, the anomalies z' were further scaled by a factor inversely proportional to the sine of latitude, ($\sin 45^{o} / \sin \theta$). As discussed in DG, this scaling, which is similar to that used when obtaining a geostrophic streamfunction from height data, was motivated by studies indicating that quantities like streamfunction provided better indicators than the height fields of atmospheric energy dispersion (e.g., Hoskins *et al.*, 1977). In parts I and II, we will focus on results obtained for a set of values of the selection criteria (M , T), (+100 m, 10 days) and (-100 m, 10 days), that appears broadly representative of a large range of values.

The data were low-pass filtered *before* applying the selection criteria (using the low-pass filter described in Blackmon, 1976) in order to remove possible effects of brief transient fluctuations on the starting (and ending) times. Case dates were determined from time series at certain "key" points (45^{o} N 170^{o} W for the Pacific cases and 50^{o} N 25^{o} W for the Atlantic cases) that were previously identified as the locations of the regional maxima in the total numbers of (positive and negative) persistent anomaly cases (DG) .

Composite time evolutions for cases of a given sign and region were then obtained by averaging over all the selected cases relative to the development (or breakdown) times to obtain the:

1) "low-pass" filtered evolution (periods > 10 days) at 500 mb;
2) corresponding "unfiltered"time evolution at 500 mb;
3) evolution of anomaly vertical structure at all levels from the surface to 100 mb;
4) evolution of full fields (i.e., mean plus anomaly) to describe and interpret associated synoptic features;
5) budgets of various quantities (e.g., of heat and vorticity) evaluated over the periods of most rapid development.

Finally, the composite results were compared with the results obtained from analyses of individual cases.

4. RESULTS

4.1. Height and zonal wind anomaly analyses

As many of the height and zonal wind anomaly results have already been presented in detail elsewhere (Dole, 1986b, hereafter LC), we will only present a few analyses to illustrate the main findings.

Figure 2 (from LC) presents composite low-pass filtered 500 mb height anomaly maps and corresponding confidence levels of the composite anomalies (estimated by a two-sided t-test with a null hypothesis of zero mean) for 14 PAC negative anomaly cases at 2-day intervals from 4 days before onset (day -4) to 6 days after onset (day +6). At day -4, the largest area of significant anomalies is located upstream over the Asian continent, with negative height anomalies exceeding the 99% confidence level over a large region extending from the Tibetan plateau eastward to Japan. Through day -2, the significant height anomalies are mainly confined to this region, with height anomalies over the central North Pacific north of 20^{o} N not significantly different from zero. The structure of the anomaly patterns prior to (and to a lesser extent following) development suggests that the associated wind anomalies are primarily in the zonal flow.

A single major negative anomaly center becomes established over the key region at day 0 and, subsequently, anomaly centers intensify in sequence downstream and to the south of the key region. Intensification of these centers occurs with little evidence of phase propagation. The gross features of the downstream developments are qualitatively reminiscent of the behavior seen in simple models of energy dispersion on a sphere away from a localized, transient (e.g., switch-on) source of vorticity (Hoskins *et al.*, 1977). A very similar pattern of behavior has also been observed by Blackmon *et al.* (1984) for time-filtered data at "intermediate" time scales (periods in the range of from 10 to 30 days), who also interpret

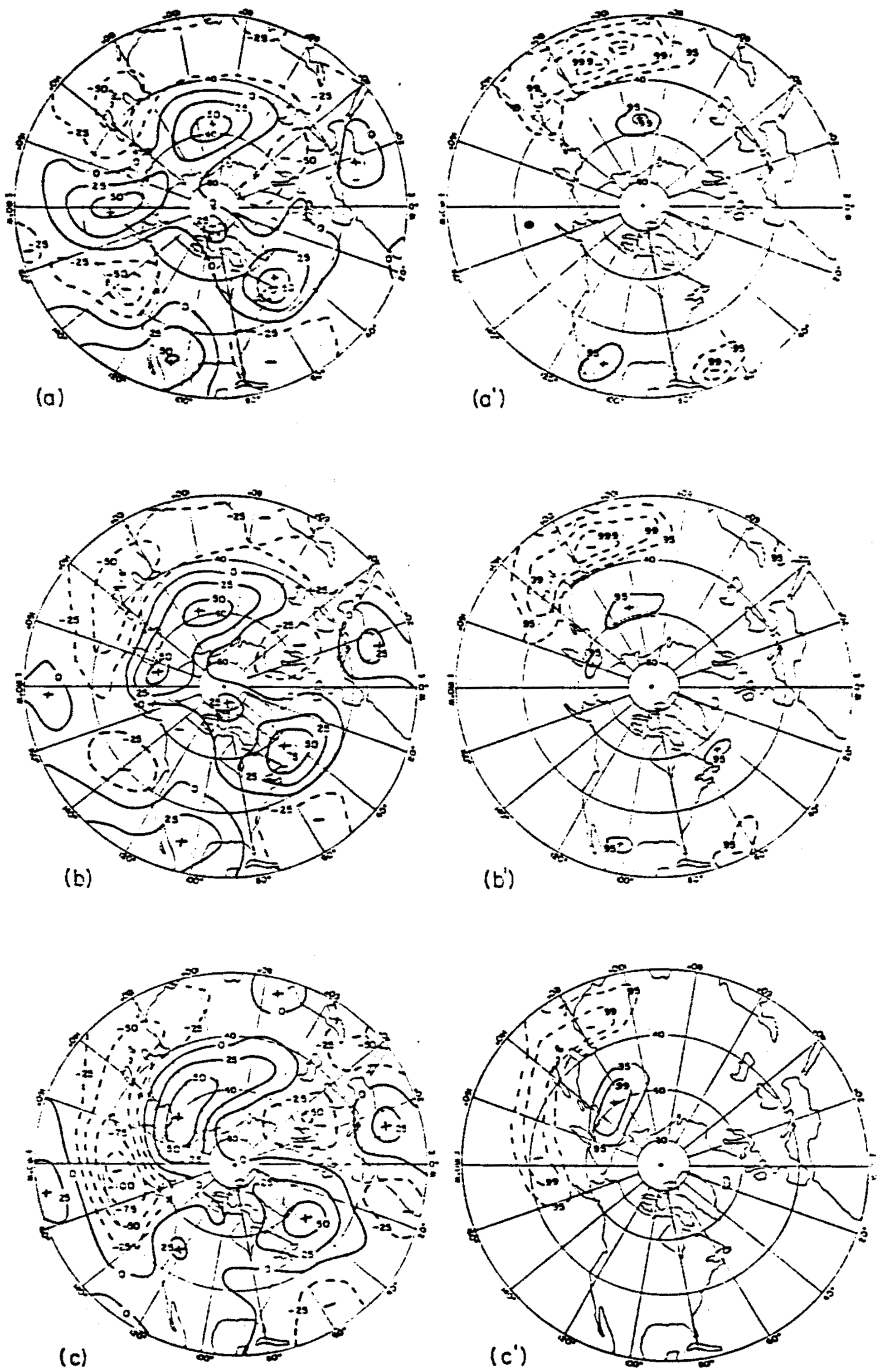


Fig. 2. Composite evolution of low-pass filtered 500 mb height anomalies (units: m) obtained from 14 PAC negative anomaly cases for days (a) -4 ; (b) -2 ; (c) 0 ; (d) +2; (e) +4; and (f) +6 relative to onset time. Areas where the composite mean anomalies are greater or less than zero at the 95%, 99% and 99.9% levels are shown for the same times in (a') - (f'). Dashed lines correspond to negative anomaly values. Solid dot in (a') denotes the location of the key point used in constructing composites.

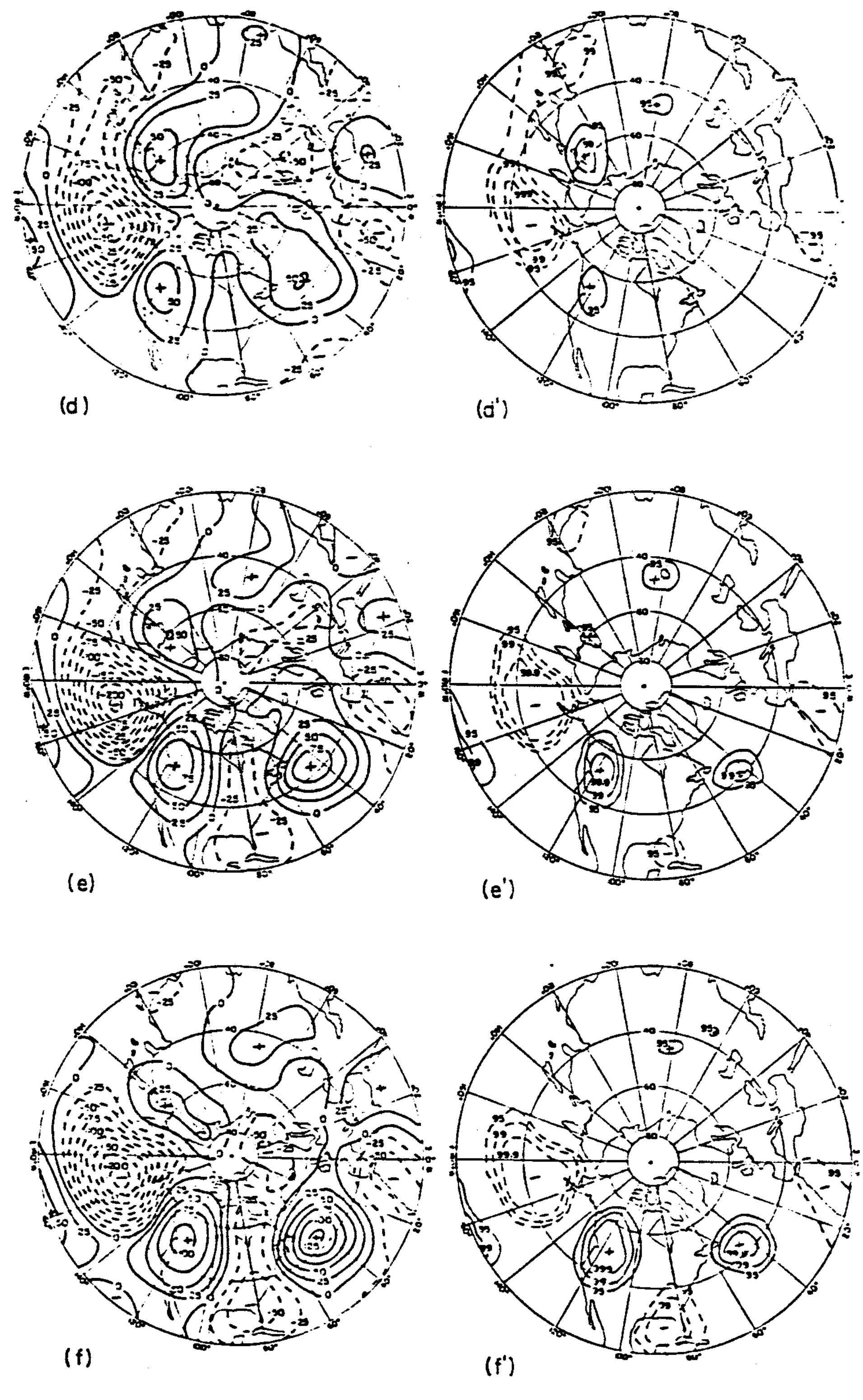


Fig. 2 (cont.)

their results for these periods as predominantly a manifestation of horizontal energy dispersion by Rossby waves.

Similar sequences for the positive anomaly cases (not shown, see LC) display a number of striking similarities, with the exception of the intense positive anomaly center that develops over the western Atlantic in the negative cases which has no clear counterpart in the positive cases. Once established, the large-scale anomaly pattern strongly resembles the Pacific - North American (PNA) teleconnection pattern (Wallace and Gutzler, 1981).

Although data used in the previous analyses were low-pass filtered, the main center over the Pacific appeared to develop rather rapidly. In order to obtain further clues as to the nature of this rapid development, identical analyses were conducted instead on unfiltered data.

Figure 3 displays the unfiltered 500 mb anomaly composite evolution and the corresponding confidence levels for the same cases at 1 - day intervals from day -3 to day 0 . The major *additional* feature not evident in the low-pass filtered analyses is associated with a synoptic-scale disturbance located to the east of Japan at day - 3. This disturbance propagates eastward while intensifying. Intensification continues as the center approaches the key region by around day 0 , although from this time onward little further propagation is evident (not shown). A second negative anomaly center located over the East China Sea on day - 2 follows a similar course, eventually merging with the main center after day 0. As in the filtered analyses, throughout this time a larger scale pattern of height falls over the North Pacific to the north of the jet axis with height rises to the south is also evident. Maps following day 0 are not shown but appear qualitatively similar to those displayed earlier. In addition to the jet variations identified previously, then, the unfiltered analyses suggest that the initial rapid growth of the main center may also be associated with an eastward-propagating synoptic-scale disturbance of apparently mid-latitude origins.

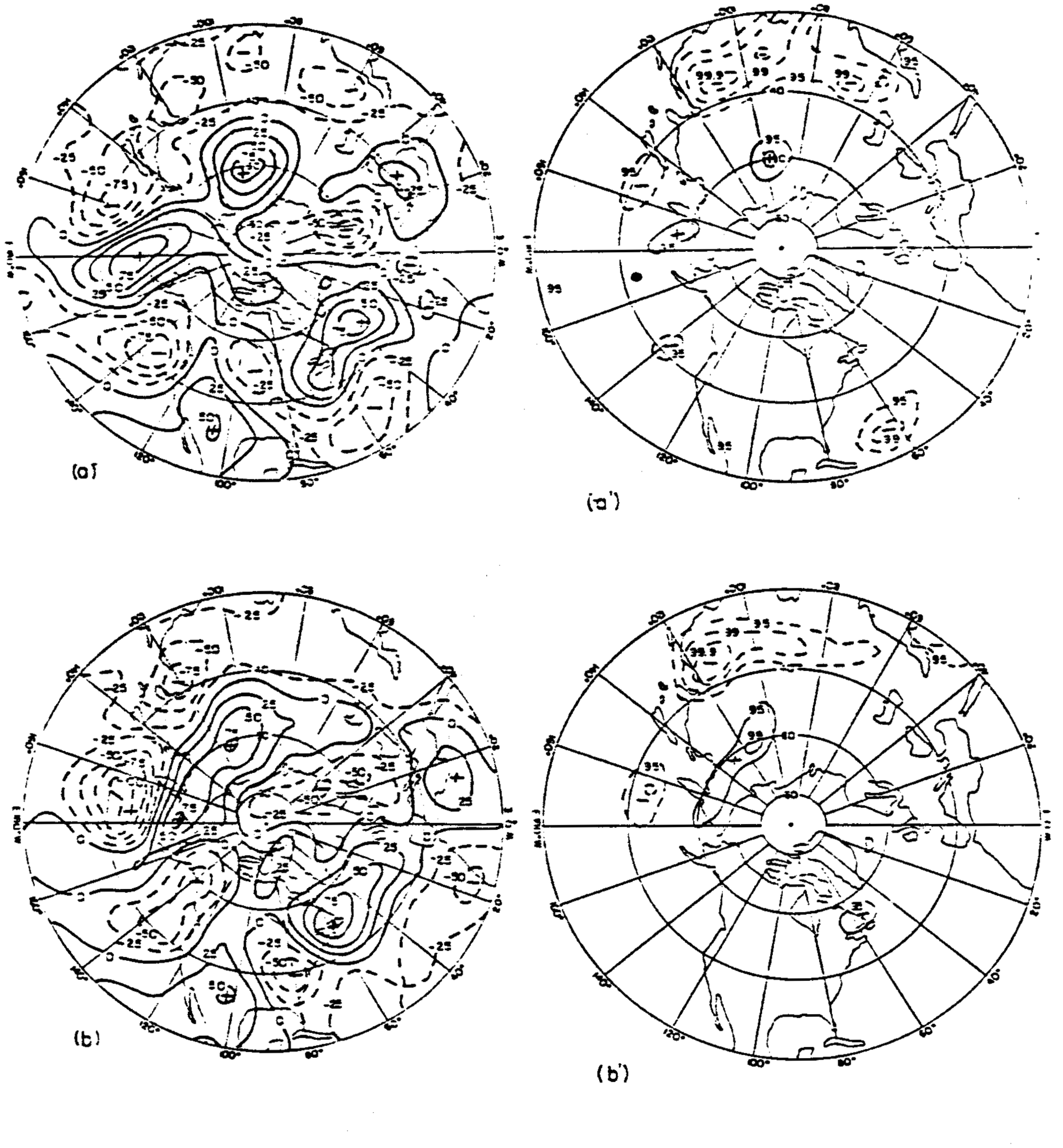


Fig.3. Composite evolution of unfiltered 500 mb height anomalies (units: m) obtained from the same PAC negative anomaly cases as in Fig. 2, for day (a) - 3, (b) -2, (c) -1, and (d) 0 relative to onset time. Corresponding confidence levels are shown in (a') - (d').

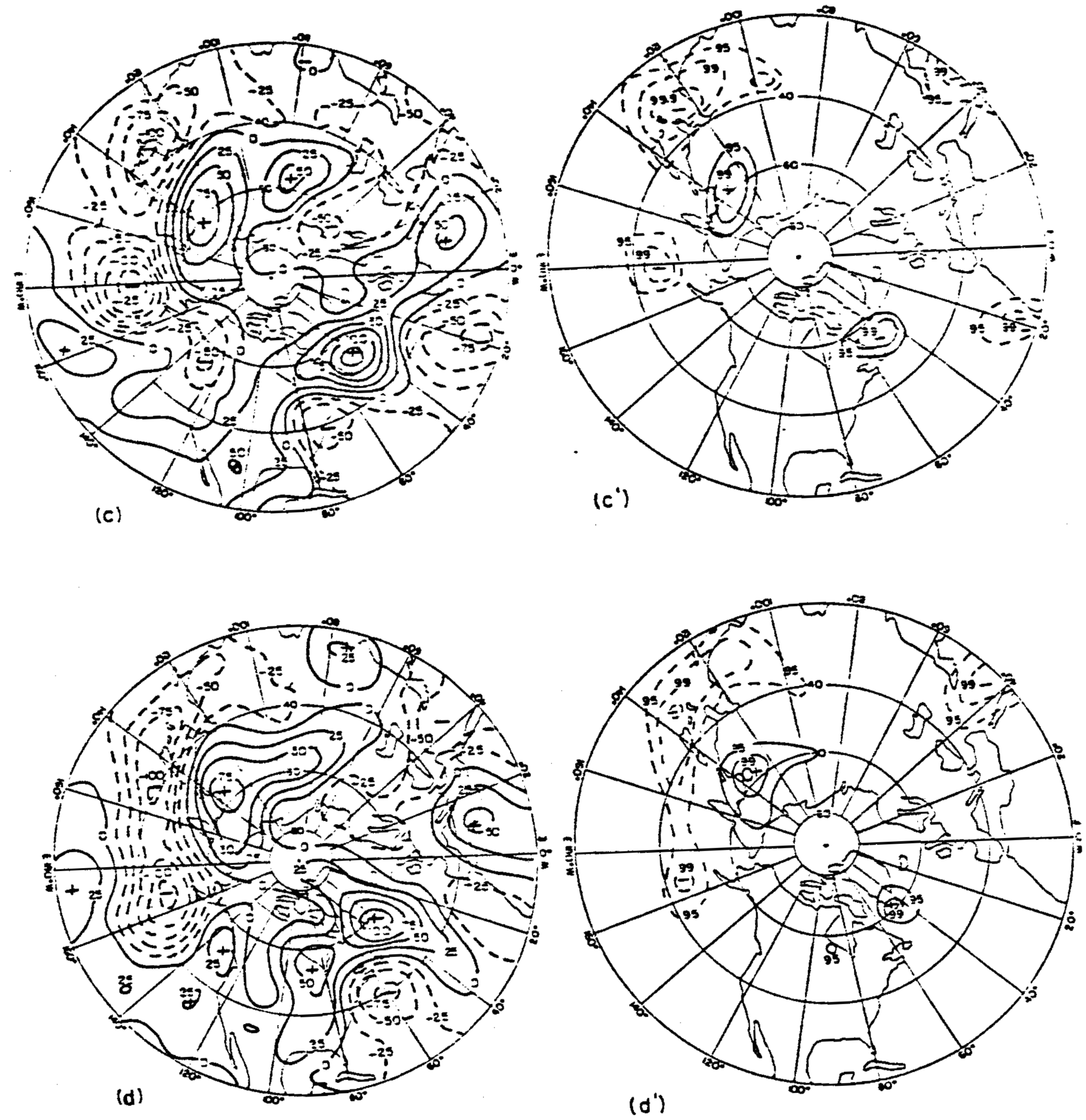


Fig. 3 (cont.)

The primary results to emerge from both filtered and unfiltered 500 mb analyses (LC) are:

1) The development rates are often rapid (order of a week), with no evidence of a statistically significant precursor over the central North Pacific until just prior to day 0 ;

2) Intensification downstream from the main center occurs with little evidence of phase propagation, qualitatively resembling the behavior seen in simple models of horizontal energy dispersion on a sphere away from a transient, localized source (e.g., Hoskins *et al.*, 1977).

3) There are at least two possible precursors to the developments, one related to variations in the jet intensity and structure over eastern Asia and the southwestern North Pacific and the other to an eastward - propagating, intensifying synoptic-scale disturbance located over eastern Asia at day - 5 .

In order to clarify the roles played by these features in the persistent anomaly developments, we will first consider the variations in the zonal flow over eastern Asia and the western Pacific, and subsequently examine the evolution of the synoptic-scale disturbance. All remaining analyses were performed on unfiltered data.

We note from Fig. 2 that the evolving height anomalies show an interesting relationship to the climatological mean jet structure, with a tendency for the largest meridional gradients in the height anomalies to occur over the mean jet axis. The implied intensification and eastward extension of the jet can be seen more clearly in Fig. 4 (from LC), which displays the time evolution of the 500 mb geostrophic wind anomalies for the PAC negative cases evaluated along the wintertime climatological mean jet axis from central Asia to the eastern North Pacific. Well prior to development, there is a modest intensification of the jet well upstream of the key region, with a weaker than normal jet to the east near the jet exit region. Beginning at about day - 3 , the zonal wind anomalies in the western Pacific strengthen and begin to propagate eastward, reflecting an intensification and eastward shift of the jet

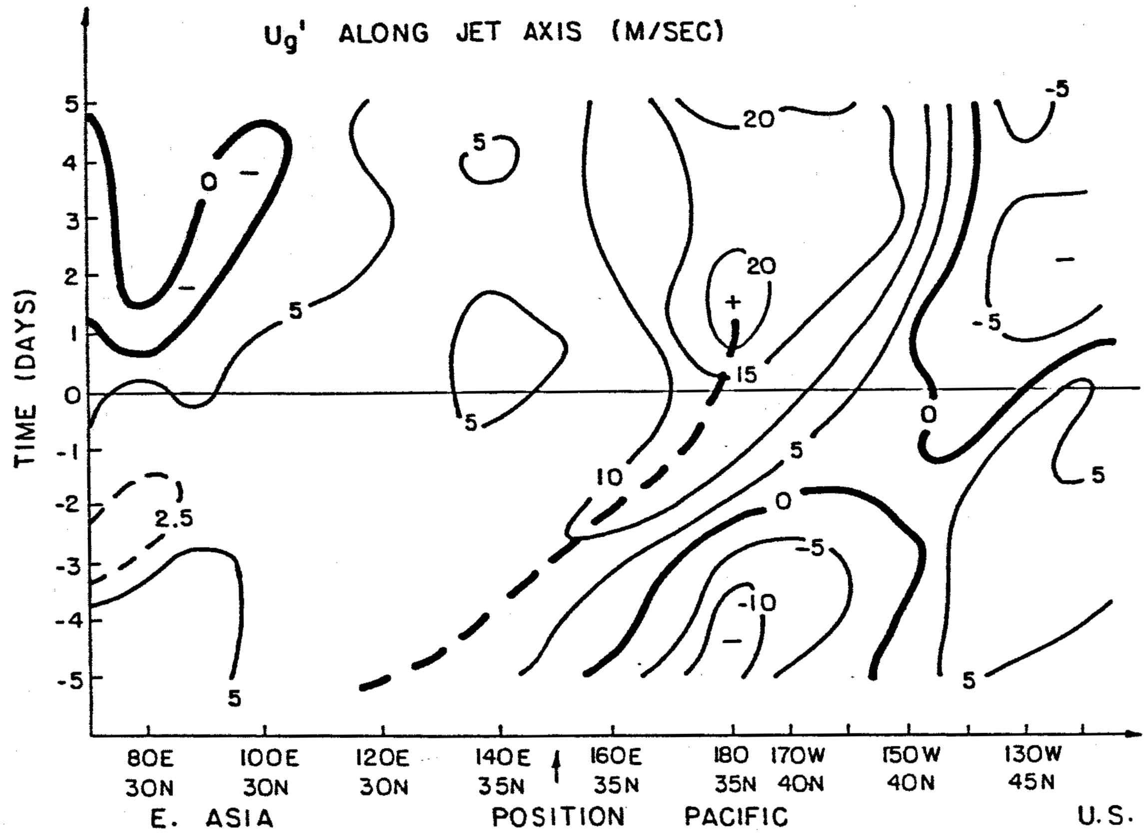


Fig. 4. Time evolution of the composite geostrophic wind anomalies (in m s^{-1}) evaluated along the wintertime climatological mean 500 mb jet axis from eastern Asia to western North America for the period from 5 days before to 5 days following onset of the PAC negative cases. Arrow on abscissa denotes the location of the climatological mean jet maximum.

maxima. A significant intensification of the anomalies is evident as they propagate eastward of the jet exit region located near 160° E. Following development, the major flow anomalies remain nearly quasi-stationary over the Pacific, reflecting the intensified and eastward-elongated jet structure.

Parallel analyses on the positive cases (not shown, see LC) show instead a modest weakening of the jet upstream of the key region prior to development, followed by a pronounced weakening of the jet in the exit region over the central North Pacific, with a tendency to form a second jet well to the north of its climatological mean position. Analyses conducted on the Atlantic cases (not shown) also display similar variations in the jet intensity and structure, with a tendency for a weakening of the upstream jet over eastern North America and the western Atlantic prior to the development of the positive cases (which are associated with blocking over the eastern Atlantic) and an intensification of the western Atlantic jet prior to the development of the negative cases.

LC presents a cross-sectional analysis (not shown) indicating that the evolving jet located over eastern Asia and the western Pacific prior to development is associated with pronounced thermal anomalies, with anomalously cold air centered at about 40° N. The thermal structure in this region is associated with an upward increase in positive vorticity, with maximum positive vorticity anomalies located near the tropopause at about 40° N. In the presence of mean westerly winds, we anticipate through quasi-geostrophic theory (e.g., Holton, 1979) that downstream of the vorticity maxima there will be an anomalous upward increase in positive vorticity advection with an enhanced likelihood for strong cyclogenesis.

4.2. Barotropic conversions during the developments

In both the Pacific and Atlantic cases, although zonal wind anomalies are first observed upstream of the climatological mean jet maxima, prominent intensification occurs following the time when the anomaly maxima propagate downstream into the jet exit regions over the central oceans. This behavior resembles the result obtained by Simmons *et al.* (1983), who found that zonally-elongated eddies (i.e., having u'^2 greater than v'^2) that are located in regions where the time-mean zonal flow is decreasing downstream are capable of growing barotropically by extracting kinetic energy from the time-mean flow. In particular, the patterns that we observe bear a marked similarity to the structure of the most unstable mode of the January climatological mean 300 mb flow obtained by Simmons *et al.*, suggesting that this mechanism may contribute positively to the developments.

Following Hoskins *et al.* (1983) and Simmons *et al.*, we have obtained local estimates of the rate of increase of kinetic energy of the anomalies $K' = (u'^2 + v'^2)/2$ due to barotropic conversions C from the time-mean flow u into the eddies from

$$C \equiv E \bullet \nabla \bar{u} \qquad (1)$$

where the components of the "E - vector" (really a pseudovector) are defined by

$$E = (v'^2 - u'^2 , - u'v') \qquad (2)$$

and the primed quantities denote departures from the climatological-mean wintertime values.

Daily and time-average E - vectors and corresponding local barotropic conversion estimates have been obtained by two methods: first, by evaluating u'^2, v'^2, and $u'v'$ and associated daily (and time- average conversions) directly from the composite anomaly evolution (the "E - vectors of the composite") ; and, second, by evaluating u'^2, v'^2, and $u'v'$ and associated daily conversions directly from the case anomaly evolutions for each case and then averaging over all of the cases (the "composite E - vectors"). That is, the E -vectors of the composite are given by

$$E_1(t) = \left(\langle v' \rangle^2 - \langle u' \rangle^2 \, , \; - \langle u' \rangle \langle v' \rangle \right) \qquad (3)$$

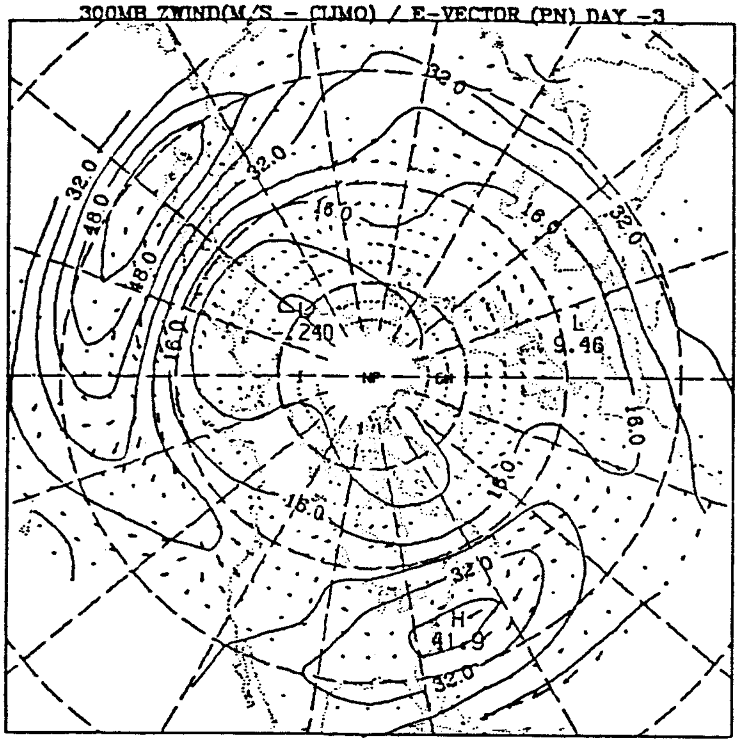
while the corresponding composite E - vectors are obtained from

$$E_2(t) = \left(\langle v'^2 - u'^2 \rangle , \; - \langle u' v' \rangle \right) \qquad (4)$$

where the angle brackets denote an ensemble average over all of the cases at time t in the evolution. The first approach essentially assumes that the composite evolution represents the features of interest, while the second approach (which includes contributions from case-to-case variability) provides a more direct basis for comparisons with climatological mean estimates of local barotropic conversions that are presented elsewhere (Wallace and Lau, 1986). There are a number of rather subtle aspects involved in interpreting local energy conversions that will not discussed here, as our objective is to describe certain characteristic aspects of the developments; however, the interested reader may consult Simmons *et al.* for a more thorough discussion.

The basic features obtained from the E - vector analyses are illustrated in Figures 5a and 5b, which for 300 mb level data display, respectively, the E -vectors of the composite, $E_1(t)$, on days - 3 and + 3 superposed over the climatological -mean values of the zonal flow. Corresponding maps of composite E - vectors , $E_2(t)$, are not shown, but are qualitatively similar, although with a tendency toward a more eastward orientation, presumably due mainly to additional contributions from synoptic-scale disturbances that are

(a)



(b)

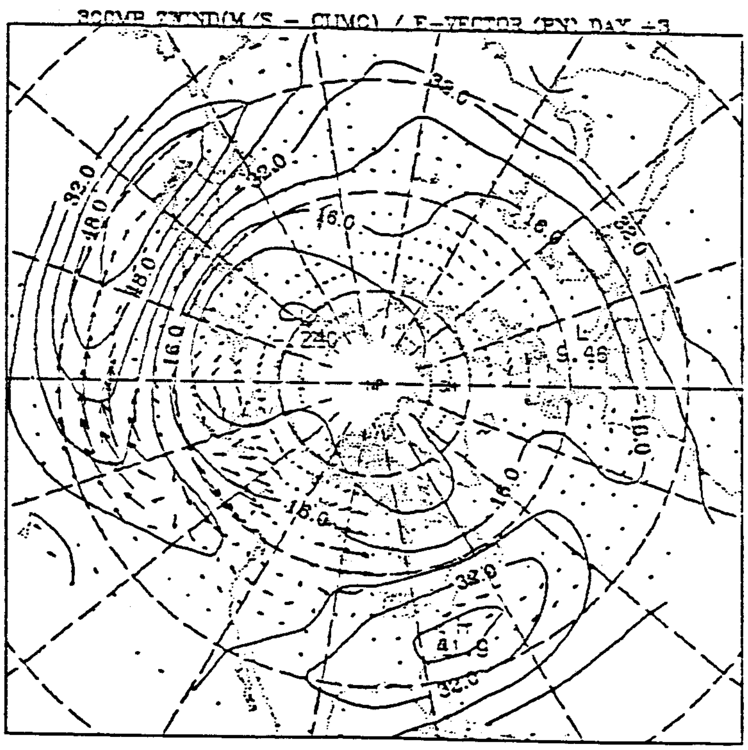


Fig. 5. E vectors of the composite obtained from unfiltered 300 mb data on days (a) - 3 and (b) + 3. Also shown are isotachs of the wintertime climatological mean 300 mb zonal flow (units: m s^{-1}) for the14 winter seasons. The largest E vector has a magnitude of 615 m^2s^{-2} (located in (b) near 40^o N 165^oW). For reference, the corresponding wintertime climatological maxima over the North Pacific is about 430 m^2s^{-2}.

removed in the compositing procedure. In brief, the basic results of the E-vector and local barotropic conversion estimate analyses indicate that:

1) initially (Fig. 5a), the pattern of **E** -vectors is rather poorly organized over the Pacific. Following day 0, however, the pattern becomes distinctly more organized, with a clear up-gradient orientation evident over the North Pacific downstream of the climatological mean jet maxima (Fig. 5b). From (1), we see that the structure of the disturbance has therefore become favorable for positive local barotropic conversions from the time-mean flow.

2) More detailed calculations of the local barotropic conversions (not shown) indicate that, until about day 0, local barotropic conversions from the mean flow into the eddies are, if anything, *smaller* than climatological mean values. Following this time, the local barotropic conversions over the North Pacific considerably exceed climatological mean values.

The direction of the **E** - vectors also provides an indication of the direction of the horizontal group velocity relative to the mean flow (when that concept is applicable), with the meridional component of the **E** - vector pointing in the same direction as the direction of meridional energy propagation for Rossby waves (Hoskins *et al.*, 1983). In common with the patterns obtained throughout the developments, the **E**-vector patterns displayed in Fig. 4 provide *no* indication that enhanced energy propagation is occurring from the tropics toward mid-latitudes either immediately prior to or following the developments.

4.3. Synoptic characteristics of developments

The analyses so far have illustrated the evolution of *anomalies* in the height and zonal wind fields preceding the development of persistent negative anomalies over the North Pacific. We will now examine the corresponding synoptic features, focusing particularly on the potential relationship between the synoptic-scale disturbance identified in the unfiltered anomaly analyses and the larger scale developments.

Figure 6 displays the evolution of the composite 500 mb height patterns at 2 - day intervals from 5 days before onset until 5 days following onset. At day - 5 (Fig. 6a), a weak ridge is located over the central North Pacific, while the Aleutian low is abnormally weak and displaced well to the west of its climatological mean position. At this time, a strong short-wave trough is just approaching the coastline over extreme eastern Asia. Through day - 3 (Fig. 6b), this trough continues to propagate eastward, reaching the central Pacific to the west of the dateline. At this time, the trough axis to the north of the jet axis displays a pronounced northwest-southeast tilt, suggestive of an enhanced southward transport of westerly momentum into the latitude of the jet. Just equatorward and upstream of the short-wave trough, the strong westerly flow that is initially confined to the far western Pacific begins to elongate eastward and intensify, so that by day - 1 (Fig. 6c), the region of strong zonal flow extends over most of the western half of the Pacific. Through this time, the upper-level short-wave has continued to propagate eastward and intensify, forming a second low center over the Aleutians. Following this time, this feature continues to deepen, although little further propagation is evident. Further downstream, pronounced large-scale amplification occurs, first in the ridge near the west coast of North America (Fig. 6d) and, subsequently, the trough over eastern North America (Fig. 6e). By day + 5 (Fig. 6f), the highly amplified 500 mb height pattern that is characteristic of the time-average structure of the PAC negative anomaly cases (DS) is fully established.

Figure 7 displays the relationship between the intensifying surface low and the evolving

(a) (d)

(b) (e)

(c) (f)

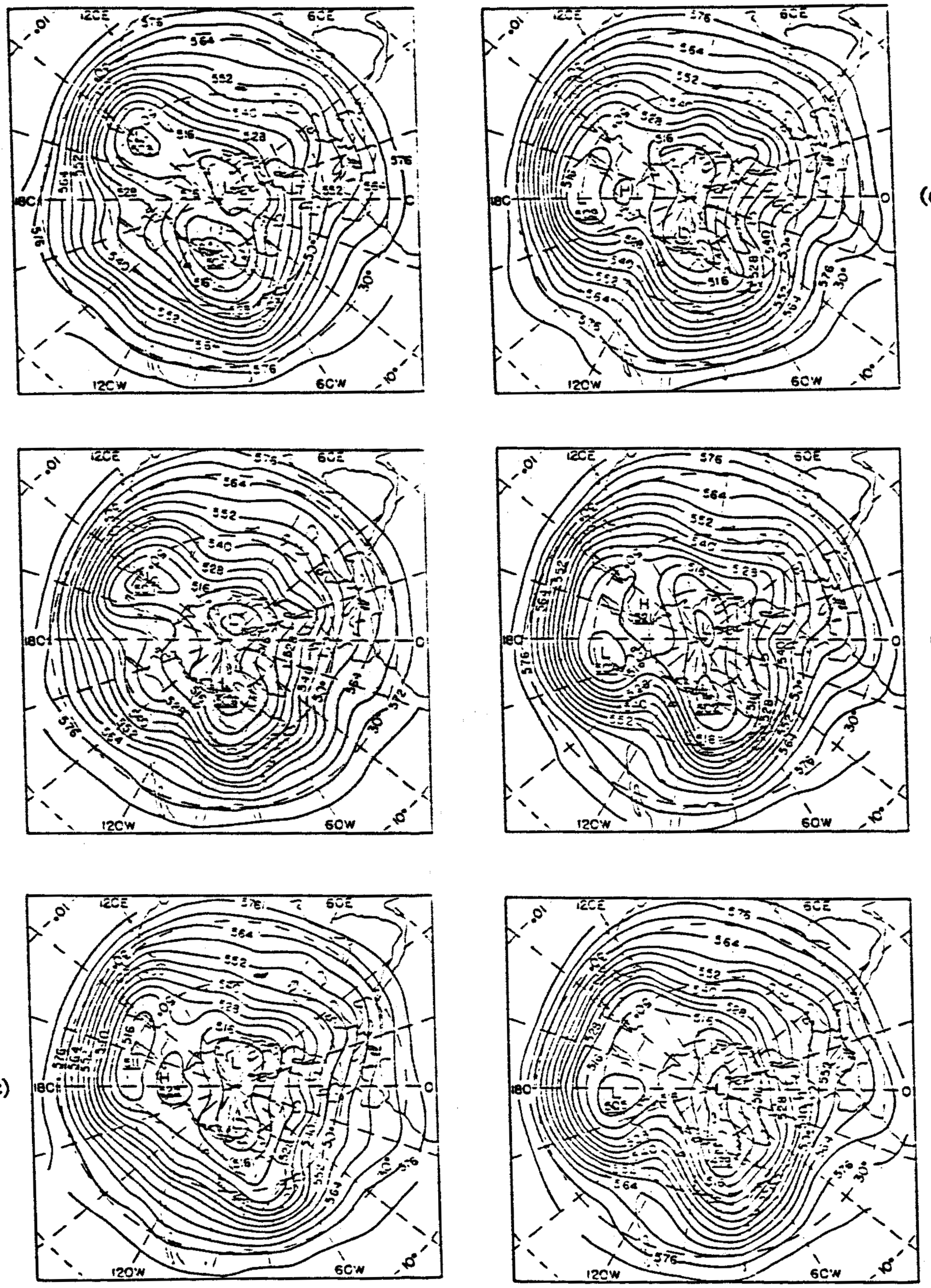


Fig. 6. Composite evolution of unfiltered 500 mb heights (units: m) for days (a) - 5 ; (b) - 3 ; (c) - 1 ; (d) + 1 ; (e) + 3 and (f) + 5 relative to onset time.

thermal structure (as reflected in the 1000 - 500-mb thickness fields) at times during the early and later stages of the development (days - 3 and + 3, respectively). The developing low to the northeast of Japan on day -3 (Fig. 7a) is located on the cyclonic shear side of the jet in a region of pronounced baroclinity, a region particularly favorable for rapid baroclinic growth. On its western flank, strong cold advection is occuring on the north side of the jet axis from off the Asian continent to over the western North Pacific. The geostrophic flow over the western Pacific appears strongly frontogenetical at this time, with the advection tending to increase the temperature gradient most rapidly near the latitude of the intensifying jet. Through this time, a number of aspects of the development are reminiscent of features commonly observed during major eastern Asian "cold surge" events (e.g., Joung and Hitchman, 1982; Boyle, 1986a,b).

Between day -3 and day +1 (not shown), the low continues to propagate eastward and intensify, forming a single major center over the central North Pacific. During the time of eastward propagation, the region of geostrophic cold advection to the north of the jet axis also extends eastward to the west of the surface low, again tending to increase temperature gradients over the western and central Pacific at latitudes near 40^{o} N. By day +1 , the developing southerly flow to the east of the major low center produces strong warm advection over the northeastern North Pacific, northwestern North America, and Alaska under and to the west of the developing upper level ridge. By day +3 (Fig. 7b), an intense cyclonic circulation encompasses most of North Pacific north of 35^{o} N . At this time, the major areas of thermal advection are mainly near the continental margins, with pronounced cold advection continuing southward and eastward out of Asia toward the western Pacific and warm advection occurring from the eastern Pacific toward Alaska.

(a)

(b)

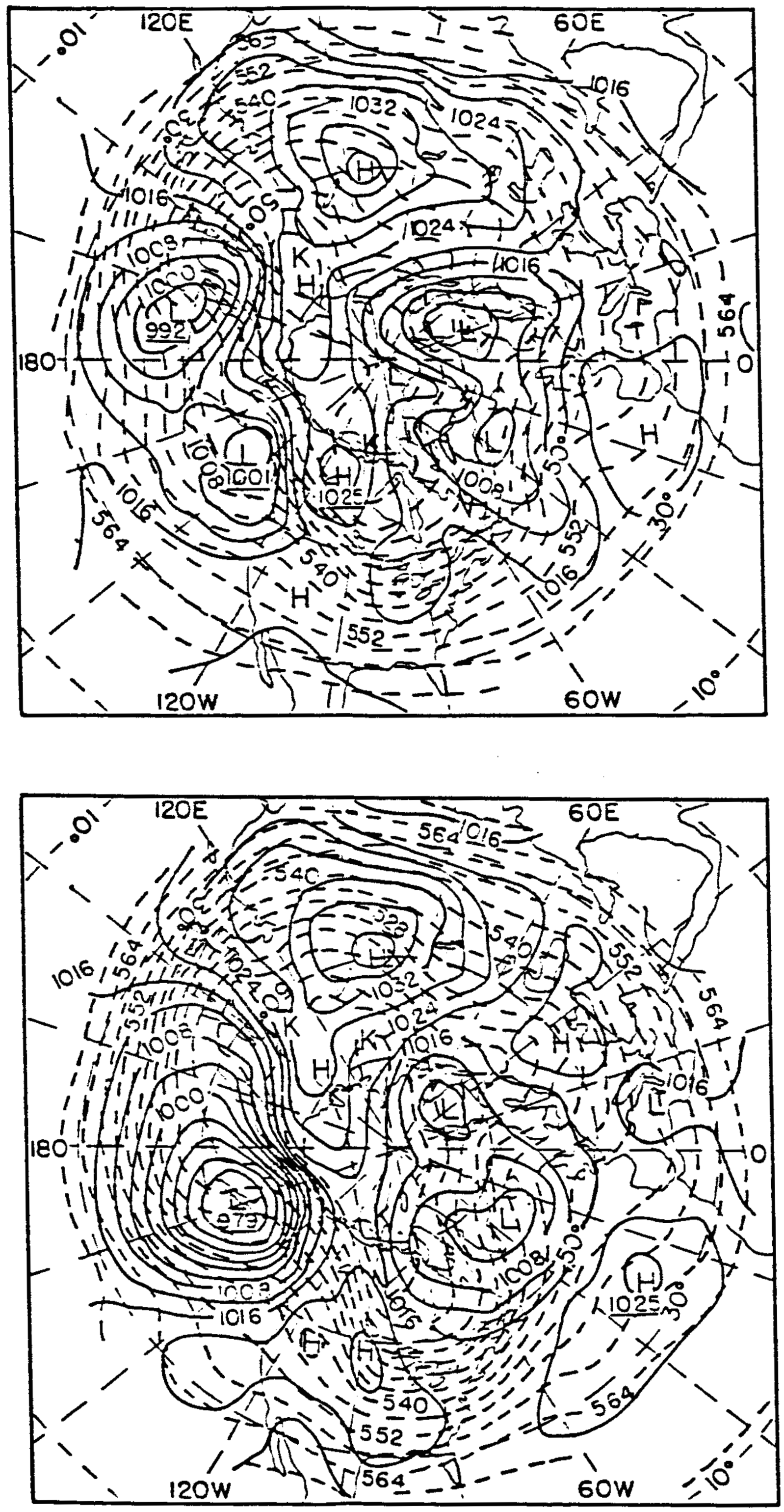


Fig. 7. Composite sea-level pressure (units: mb) and 1000 - 500 mb thickness fields (units: dam) at days (a) - 3 and (b) + 3 relative to onset time.

4.4. Q - vector and eddy heat flux analyses

Q - vector (Hoskins *et al.*, 1978; Hoskins and Pedder, 1980) and eddy heat flux vector analyses have been conducted to further examine the role of baroclinic processesses in the developments. As Hoskins and Pedder indicate, the rate of frontogenesis due to the geostrophic flow $\mathbf{V_g}$ is given by

$$\frac{D\,|\,\nabla\theta\,|^2}{Dt} = 2\,\mathbf{Q} \bullet \nabla\theta \qquad (5)$$

where $\mathbf{Q}$, equal to the rate of change of potential temperature gradient following the horizontal geostrophic velocity $\mathbf{V_g}$, has components

$$\mathbf{Q} = -\left(\frac{\partial \mathbf{V}_g}{\partial x} \bullet \nabla\theta\ ,\ \frac{\partial \mathbf{V}_g}{\partial y} \bullet \nabla\theta \right) \qquad (6)$$

As Hoskins and Pedder show, ascent is favored in regions where $\mathbf{Q}$ is convergent, with $\mathbf{Q}$ tending to point into (opposite) the direction of lower-level (upper-level) ageostrophic motion.

Fig. 8 displays **Q** - vector and temperature analyses obtained from the 700 mb composite fields at days - 3 and + 3 . Again, there are a number of interesting features that are apparent in these analyses, but two particularly significant characteristics to note are, first, that the **Q** - vectors in the western Pacific region are consistently directed toward the warmer air, indicating that the geostrophic flow in this region is frontogenetical (average of roughly double climatological values), as inferred previously, and second, that the **Q** - vectors tend to be convergent in the region of the thermal ridge and divergent in the region of the thermal trough at this level. This suggests that at this level the net vertical eddy heat fluxes are upward ($w'T' > 0$), as would be expected for a growing baroclinic disturbance.

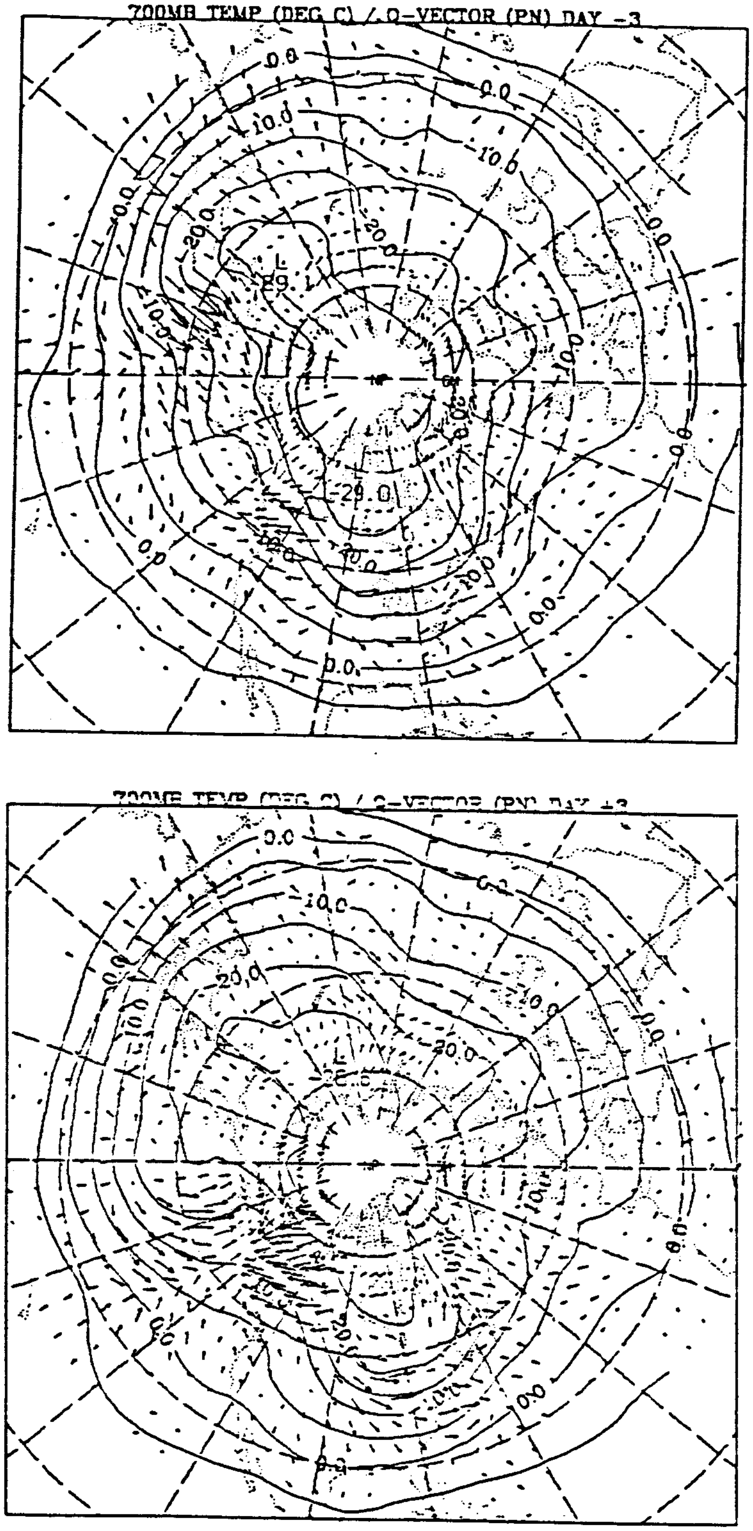


Fig. 8. Composite 700 mb temperatures (units: C) and corresponding Q vectors of the composite on days (a) - 3 and (b) + 3 relative to onset time.

We might anticipate from the overall structures of the patterns that horizontal heat flux statistics obtained through this period would show significant correlations in the zonal ($u'T'$) as well as meridional ($v'T'$) fluxes, with a marked tendency for the heat flux vector to be directed down the climatological mean temperature gradient. That this is the case can be seen in Fig. 9, which displays for 700 mb data the average heat fluxes of the composite, $\langle V' \rangle \langle T' \rangle$, superposed on the climatological mean temperature field for averages over days (- 5 , 0), (1 , 5) and (- 5 , 5) . We see that, for all of the averaging periods, the net fluxes associated with the composite structure are down the climatological mean temperature gradient , although most consistently so during the first part of the development. Following day + 1 , the fluxes also display evidence of a non-divergent (rotational) component over the central Pacific, consistent with the more equivalent barotropic structure of the disturbance observed at later stages of the development. Even during this period, however, the net heat fluxes are directed toward lower mean temperatures. Major contributions to the downgradient heat flux shift from the western Pacific early in the period to the eastern Pacific at later times. Composite heat fluxes $\langle V' \; T' \rangle$ (obtained instead by determining the fluxes on each day for each case and then averaging) typically have considerably larger magnitudes and are almost exclusively downgradient, presumably due to the additional contributions from mobile baroclinic disturbances that have been removed by the compositing procedure.

The overall impression that emerges from a more detailed inspection of the daily maps and heat flux analyses is one of initial baroclinic growth at synoptic-scales, evolution toward a more equivalent barotropic structure (corresponding with reduced vertical tilts of the disturbance, weak horizontal temperature advections and small downgradient heat fluxes) with increasing barotropic energy conversions by around day 0, with possible additional baroclinic contributions occurring again later in the development . At the later time, the growth appears to occur on a scale that is larger than typical for synoptic-scale cyclogenesis.

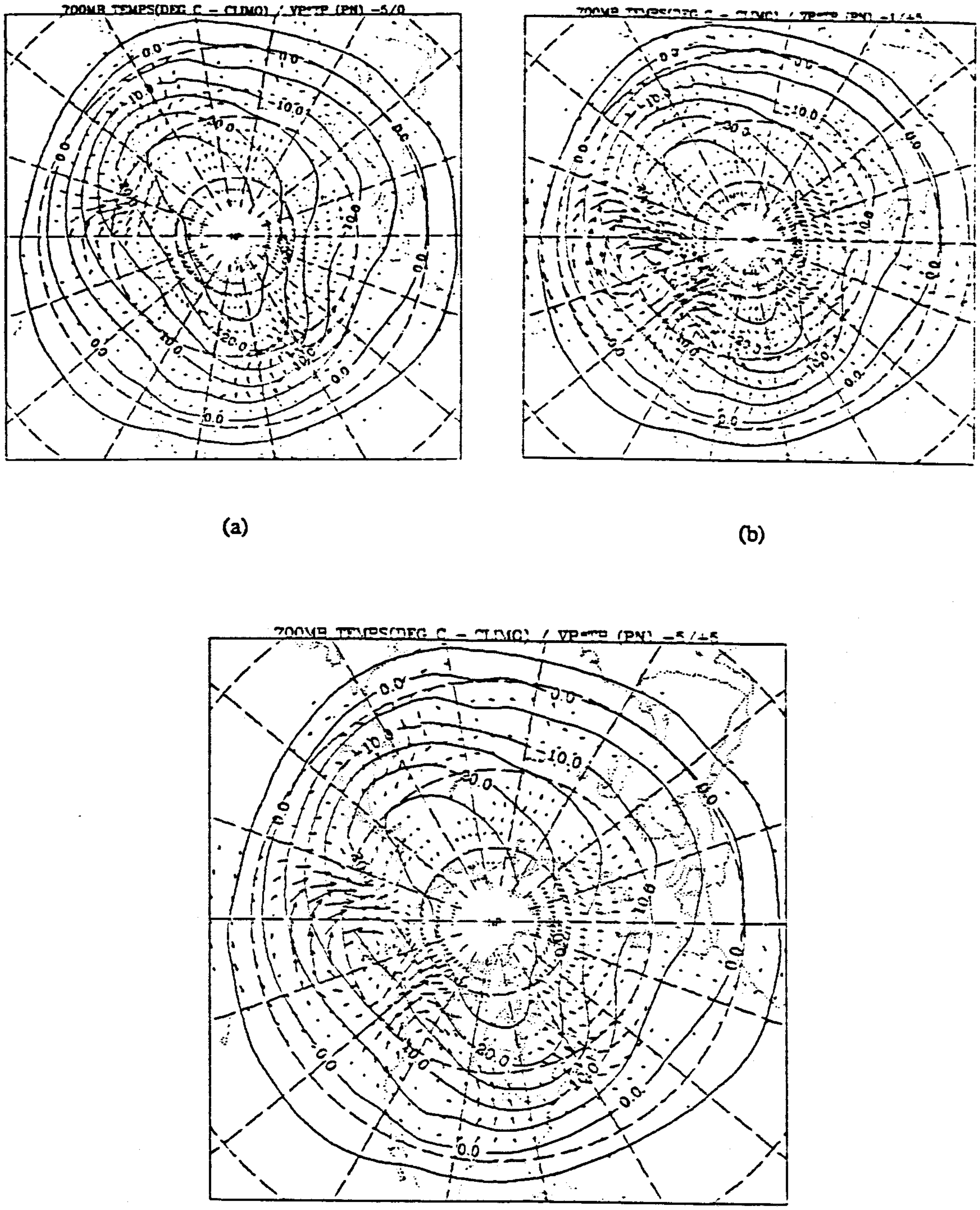


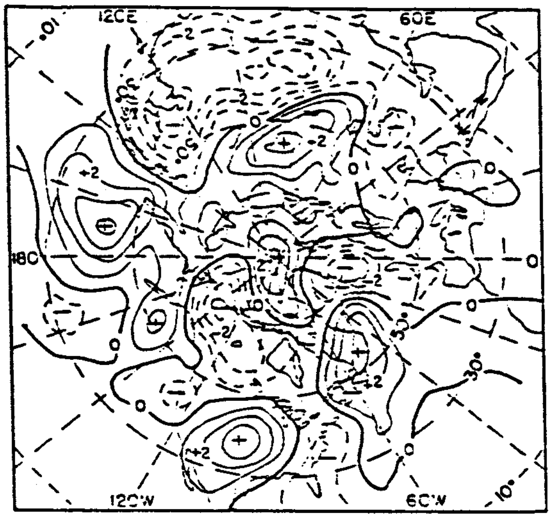
Fig. 9. Time average heat fluxes of the composite anomalies ⟨ v ' ⟩ ⟨ T ' ⟩ superimposed on wintertime climatological mean temperatures (units: C) for averages over days (a) - 5 to 0 ; (b) + 1 to + 5; and (c) - 5 to + 5. Maximum heat flux magnitudes for all periods are located over the west- central North Pacific with a maximum value of about 30 K ms^{-1} in (b).

4.5. Heat budget analyses

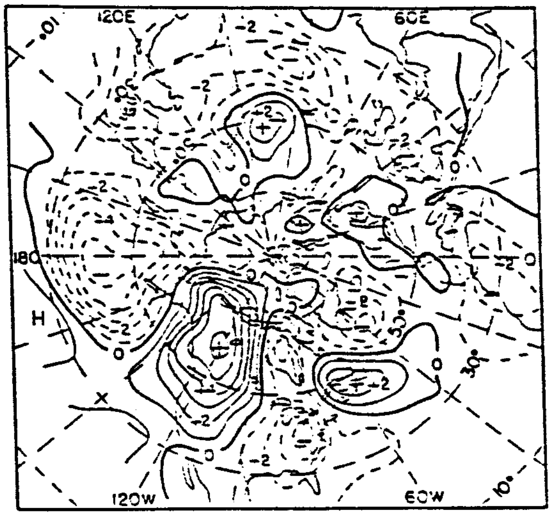
From the previous analyses, it is apparent that the large-scale temperature field has also undergone substantial changes during the period of most rapid development. These changes can be seen more clearly in Figure 10, which displays the 700 mb composite temperature anomalies at days -5 and +5 along with the corresponding composite temperature changes over this period. As inferred previously, prior to development (Fig. 10a), there is a large area of (highly statistically significant) negative temperature anomalies that at this time covers much of central and eastern Asia centered near 40° N. The negative anomalies subsequently extend eastward, covering much of the western and central North Pacific following development (Fig. 10b). The largest areas of changes are associated with the cooling over this region and with the major warming that occurs over much of the northeast Pacific, western Canada and Alaska extending into the Arctic (Fig. 10c). In addition, net cooling over the period occurs over much of the central and eastern United States, where synoptic experience suggests that the development of this pattern is often also accompanied by a major cold-air outbreak.

To obtain more quantitative estimates of the roles of various processes in producing the observed temperature changes, we have conducted heat budget analyses centered around the times of most rapid development. Assuming that the climatological mean temperature tendency is small and that variations in time of the static stability σ can be neglected, an equation for the temperature anomaly tendency may be written as:

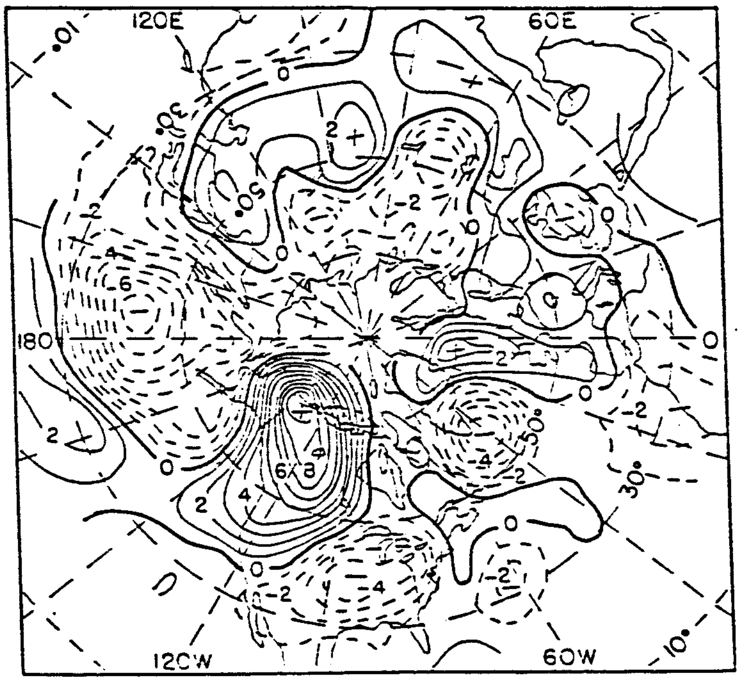
$$\frac{\partial T'}{\partial t} = -V' \bullet \nabla \overline{T} - \overline{V} \bullet \nabla T' - (V' \bullet \nabla T' - \overline{V' \bullet \nabla T'}) + \sigma \omega' + \frac{Q'}{c_p} \qquad (7)$$



(a)



(b)



(c)

Fig. 10. Composite 700 mb temperature anomalies (units: C) at days (a) - 5 and (b) + 5 , and (c) composite 700 mb temperature change between day - 5 and day + 5 (units: C).

where the overbar represents a climatological mean and the primes denote deviations from the climatological mean.The first two linear terms on the right-hand side of (7) are, respectively, the advection of the mean temperature by the anomalous winds and the advection of the temperature anomalies by the climatological-mean flow.

The non-linear term in parentheses represents the effect of deviations above or below climatological-mean values of the advection of temperature anomalies by the anomalous winds. Taken together, the first three terms represent the net effects of anomalous horizontal advection on the temperature tendency. The last two terms represent, respectively, additional contributions to the changes that result from adiabatic cooling due to anomalous vertical motions and from anomalous diabatic heating.

Heat budget calculations have been performed by averaging the terms in (7) over all cases for the period from days - 5 to + 5 in the developments. We will not present here our detailed results, but will instead cite only a few of the more important features. In brief, the results of the heat budget analyses indicate that, over most of the region of development (roughly 140^{o} E to 120^{o} W), the temperature changes over the period are positively correlated with the anomalous horizontal temperature advection and are negatively correlated with the adiabatic cooling term . These relationships are consistent with the kind of balance seen in a growing baroclinic wave, where on average air parcel trajectories must lie between level and isentropic surfaces in order to convert available potential energy (Green, 1960), but differ qualitatively from the kind of thermodynamic balance obtained in equivalent barotropic models (Charney and Eliassen, 1949), where temperature (thickness) changes are accomplished through adiabatic cooling accompanying vertical motions. Another interesting aspect of the heat budget results is illustrated in Fig. 11, which displays the net contributions to temperature changes over the period due to advection by the zonal (Fig. 11a) and meridional (Fig. 11b) winds. We see that contributions by the zonal winds and meridional winds have comparable magnitudes, and that, particularly over the western Pacific, the net advections are acting to increase the temperature gradients along the evolving jet axis.

(a)

(b)

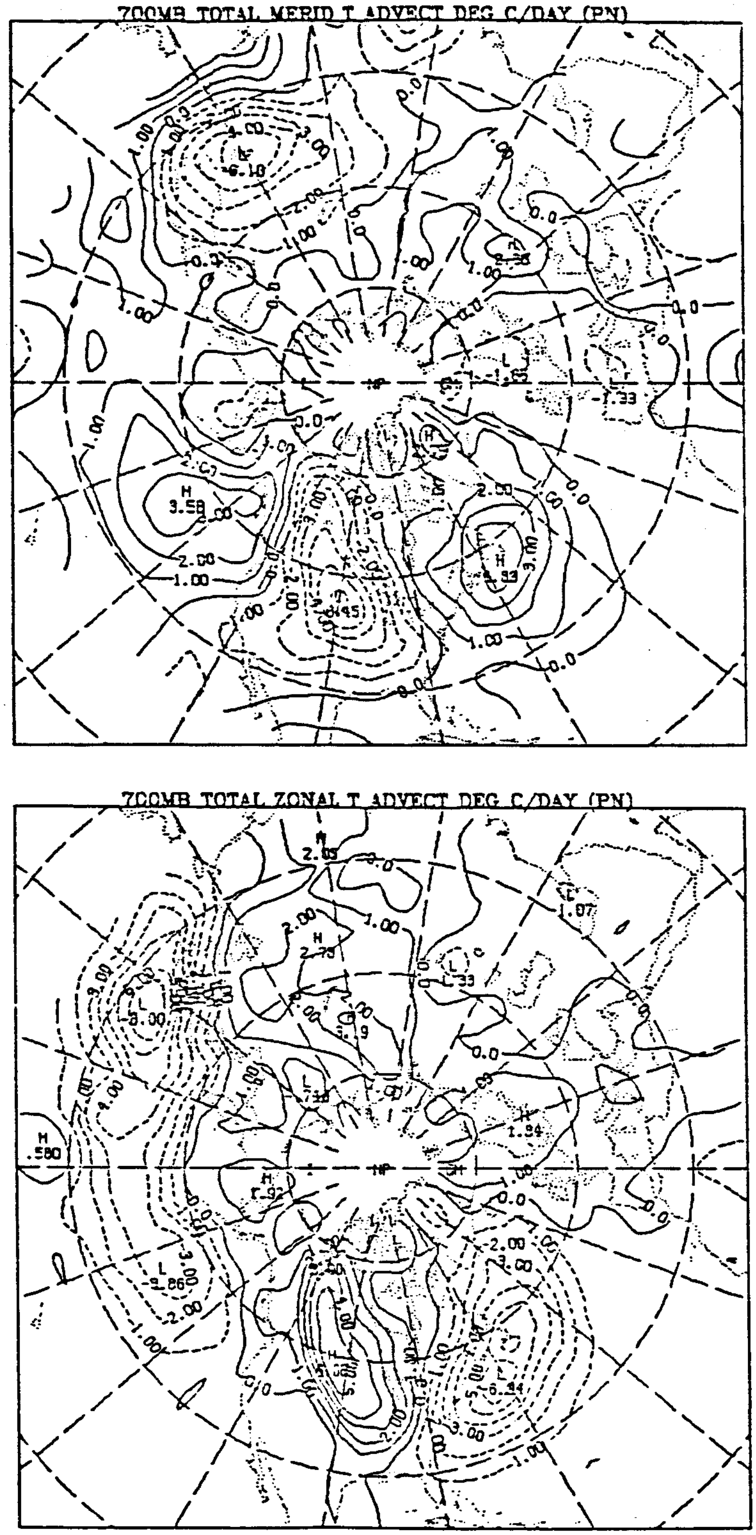


Fig. 11. Average rate of temperature change (C day^{-1}) over the period between days - 5 and + 5 due to the advection of the composite temperatures by the composite (a) meridional flow and (b) zonal flow.

5. DISCUSSION

We have found that a substantial part of the growth (and also decay) of persistent anomalies often occurs over time scales that are short compared to the time scales associated with major changes in the boundary conditions (e.g., of sea surface temperature anomalies). In addition, the sequence of development for the PAC cases suggests that the initial rapid growth of the main center may be partly associated with an eastward- propagating, mid-latitude synoptic-scale disturbance. These observations suggest that the patterns often, and perhaps primarily, grow and decay while the external forcing remains nearly fixed. General circulation modeling experiments (e.g., Lau, 1981; Blackmon *et al.*, 1986) provide further evidence that anomalies in external forcing are not required to produce patterns similar to those described here.

More detailed statistical analyses (DS) indicate that much of the temporal variability of the persistent anomaly patterns is contributed by low-frequency intra-seasonal fluctuations having integral time scales of about two weeks. Other studies of major persistent phenomena such as blocking (e.g., Rex, 1950) and of variability in time-averages (e.g., Madden, 1981) also show that events that have a profound impact on monthly-mean (and sometimes seasonal-mean) anomalies often grow and decay over times of substantially less than one month. Considerable caution should therefore be used when interpreting results based on monthly (and perhaps even longer-term) averages; in particular, interpretations of time-mean anomalies as manifestations of steady or quasi-steady phenomena may often be unwarranted.

There are a number of other potentially important implications of these results, but for now, we will emphasize only the point that *internal dynamical processes are likely to play a central role in the development and evolution of many persistent flow anomalies.* A direct operational implication is that, even in forecasts of monthly mean anomalies, initial conditions are likely to be of central importance.

Nevertheless, examination of the temporal distributions of persistent anomaly cases (see DS) does suggest that, for the PAC cases, *boundary variations* associated with the El Nino - Southern Oscillation phenomenon *have some influence in determining the relative likelihood of an event of a given sign,* with a tendency for more frequent and long-lived negative PAC events to occur during El Nino winters (although occurrences of positive events during El Nino winters and negative events during non-El Nino winters are not uncommon). Horel and Wallace (1981) have also presented observational evidence of a modest relationship between the interannual variability in tropical Pacific sea surface temperature anomalies and the PNA teleconnection pattern, which closely resembles our PAC anomaly pattern. Their finding is supported by results obtained from a number of general circulation modelling experiments on the sensitivity of the mid-latitude circulation to tropical sea surface temperature anomalies (e.g., Blackmon et al., 1983; Palmer and Mansfield, 1986).

Horel and Wallace have suggested that the extratropical circulation anomalies are essentially the manifestation of a forced stationary Rossby wave response to tropical heating anomalies located over the central equatorial Pacific. Although additional tropical data would be required to adequately test whether anomalous heating in the central equatorial Pacific is important in initiating the developments of the PAC cases, the present results provide *no* indication of any systematic northward energy propagation from this region immediately prior to the developments. An intriguing clue toward a possible link, however, is provided by the height anomaly pattern that is located well upstream over Asia and the extreme western Pacific several days prior to the initial developments over the key region. These height anomalies are associated with significant variations in the subtropical jet over eastern Asia and the southwestern North Pacific. The flow anomalies in these regions may in part be a result of changes in tropical heating over the western Equatorial Pacific and Indonesian regions, areas where interannual precipitation anomalies are known to be strongly linked to different phases of the Southern Oscillation (e.g., Horel and Wallace, 1981), and where observational studies also suggest that low-frequency intraseasonal flow fluctuations are closely coupled to variations in tropical convection (Liebmann and Hartmann, 1984; Weickmann *et al.*, 1985; Lau and Phillips, 1986).

Indeed, the latter studies do show a modest correlation between variations in convection over the tropical west Pacific and subsequent flow anomalies over the central North Pacific; however, the same analyses indicate that upstream wavetrains over Eurasia frequently *precede* outbreaks of enhanced convection over the southwestern Pacific. In addition, a number of other studies have presented results indicating that the initial flow changes in the jet region are often induced by events that have mid-latitude origins (e.g., Joung and Hitchman, 1982; Lau et al., 1983; Lau and Lau, 1984; Boyle, 1986). We also note that some theoretical results indicate that zonal flow variations in the jet region may produce substantial changes in the mid-latitude response to tropical heat sources, even if the heat sources themselves are not changing substantially with time (Jacqmin and Lindzen, 1985).

A number of simple barotropic models of extratropical low-frequency variability (Simmons, 1982; Branstator, 1983; Simmons *et al.*, 1983) have also displayed a marked sensitivity to changes in the mean flow and forcing located over southeastern Asia and the southwestern Pacific. Simmons *et al.* have suggested that the large extratropical response is related to the excitation of the most rapidly growing mode associated with a barotropic instability of the zonally-varying climatological mean wintertime flow. The growing disturbance that we have described has a horizontal structure and temporal evolution that in many respects resembles that of the most rapidly growing mode. In addition, E-vector analyses confirm that, at later stages in the evolution, the conversion from mean flow to eddy kinetic energy is likely to contribute positively to the growth of the disturbances. Barotropic instability of the zonally varying time-mean flow therefore provides one possible source for the developments in these cases.

Nevertheless, the analyses indicate that the developments are also likely to be significantly influenced by baroclinic processes. Over the North Pacific, horizontal temperature advection plays a first-order role in determining the evolution of the temperature field, with the effect of the temperature advection over the western Pacific being to increase the temperature gradient along the axis of the intensifying jet. The upper-level trough propagating eastward from Asia is initially displaced well upshear of the developing surface center over the western Pacific; this upshear displacement decreases with time as the anomalies approach maximum amplitudes. The associated net anomalous eddy heat fluxes during the development are downgradient (both poleward and westward) as well as upward. In a number of basic aspects, then, the observed characteristics of the disturbance resemble those seen in classical studies of synoptic-scale cyclogenesis (e.g., Petterssen, 1955; Palmen and Newton, 1969), although during its later stages, the spatial scale of the disturbance is considerably larger and propagation speed smaller than for typical synoptic-scale developments (Blackmon *et al.*, 1977). On the basis of the observations, then, it appears likely that baroclinic energy conversions from the mean flow also contribute significantly to the growth of the disturbance.

Perhaps the most extensive attempts to analyze the possible influences of baroclinity on the development of persistent flow anomalies have been conducted by Frederiksen (1982, 1983, 1986) in a series of investigations on the instability characteristics of the three-dimensional Northern Hemisphere wintertime mean flow. Frederiksen has been able to identify certain unstable modes whose structures and evolutions resemble in many gross aspects the features we have described. He interprets the development of the PAC pattern as essentially a two-stage process that involves both barotropic and baroclinic processes, where the stages are essentially determined by variations in the basic-state stability. During the first stage in the development , baroclinic conversions dominate and the unstable modes are eastward propagating, while in the second stage, the modes are associated mainly with barotropic conversions, have larger spatial scales and are nearly stationary. The temporal variations in the vertical structure of the disturbance (i.e., initial upshear tilts that decrease with time), however, are also similar to those observed in studies of non-modal baroclinic growth (e.g., Farrell, 1984, 1985), raising the alternative possibility that transient (initial value) baroclinic growth may instead be occurring, although on a scale larger than that usually associated with baroclinic developments.

The dominance of horizontal temperature advection in determining the evolution of the temperature fields and the potential of significant baroclinic energy conversions represent possible *direct* influences of baroclinic processes on the developments; however, *indirectly*, baroclinic processes may also play an important role in the developments. We have noted that the flow over the western Pacific during the developments is strongly frontogentical. This frontogenetical flow will tend to force a thermally direct ageostrophic circulation across the jet axis, with a tendency to produce increases in low-level convergence, upper-level divergence and ascending motions to the south of the jet over the southwestern Pacific. In combination with the southward penetration of relatively cold air at low levels, these conditions are particularly favorable for triggering enhanced convection over the tropical western Pacific, as is commonly observed following major cold-air outbreaks over eastern Asia (e.g., Chang et al., 1979; Lau, 1982). The enhanced convection will in turn tend to lead to a further intensification of the westerly jet, through Coriolis torques acting on the poleward ageostrophic flow to the north of the convection region (Lau *et al.*, 1983). Indeed, it seems likely that, under certain initial conditions, the mid-latitude and tropical interactions in this region may feed back positively to produce an anomalously large intensification of the jet and subsequently, a large amplitude downstream response. The dynamics of processes occurring in this jet region and their link to subsequent mid-latitude developments, then, present a number of fascinating problems that we believe deserve special attention, particularly in future studies aimed at understanding the potential role of tropical - mid-latitude interactions in the genesis of persistent extratropical flow anomalies.

Finally, perhaps the most problematical results concern the interpretation of the role played by synoptic-scale disturbances in the intensification of the large scale flow anomalies. We have found that during the early stages of the developments, synoptic-scale cyclogenesis occurs over the western Pacific. At later stages of the development, however, intensification appears to occur on a somewhat larger scale, eventually leading to an anomalous cyclonic circulation that extends across most of the North Pacific basin at mid-latitudes. Although a separate examination by us of the evolutions of the cases (not shown) found that in *all* cases the development of persistent negative anomalies in the central North Pacific was preceded by synoptic-scale cyclogenesis in the western Pacific, we also note that cyclogenesis frequently occurs in this region without the subsequent development of persistent large-scale anomalies. Clearly, then, synoptic-scale cyclogenesis over the western Pacific is not a *sufficient* condition for the subsequent development of persistent anomalies. More research is required, however, in order to determine whether under certain conditions the synoptic-scale disturbances directly force the larger scale circulation changes or, rather, whether indirectly they trigger or catalyze the subsequent large-scale developments.

6. CONCLUSION

We summarize the main characteristics of our cases as follows:

Development rates are often rapid, with little indication of a significant anomaly over the central North Pacific until just prior to onset. The most systematic precursors are related to variations in the jet intensity and structure over eastern Asia and the southwestern North Pacific and to an eastward-propagating, intensifying synoptic-scale disturbance that appears to have mid-latitude origins. Particularly at its initial stages, this disturbance has a markedly baroclinic structure. The cases are also typically preceded by a major buildup of cold-air over eastern Asia, with some indication in the analyses that the initial developments may frequently occur in conjunction with a major eastern Asian cold surge.

Following establishment of the major anomaly center over the central North Pacific, anomaly centers develop and intensify downstream, leading after a few days to the establishment of the persistent anomaly pattern. Much of this downstream intensification occurs with little evidence of phase propagation. In a number of aspects, this behavior resembles that seen in simple time-dependent models of energy dispersion on a sphere away from a quasi-stationary, localized source of vorticity (Hoskins *et al.*, 1977), suggesting that quasi-horizontal energy dispersion is likely to account for important aspects of the *downstream* developments. More comprehensive analyses of the evolutions than have been presented here, however, indicate that in some cases, orography and vertical propagation (and, as discussed in Part II, possibly also interactions with synoptic-scale transients) are likely to play important modifying roles.

The major anomaly center that intensifies downstream from the climatological jet maxima over the central North Pacific has a horizontal structure and temporal evolution that in many respects resembles that of the most rapidly growing normal mode associated with barotropic instability of the climatological mean wintertime flow (Simmons *et al.*, 1983). **E**- vector analyses confirm that, at later stages in the evolutions, barotropic conversions are likely to contribute positively to the anomaly development. These results indicate that barotropic instability of the time mean flow provides one possible source for the developments. At the early stages of the development, however, barotropic conversions from the mean flow into the eddies are, if anything, smaller than climatological-mean values.

Evidence obtained from a variety of analyses indicates that the developments are also likely to be significantly influenced by baroclinic processes. Horizontal temperature advection plays a primary role in determining the evolution of the temperature patterns, with zonal and meridional temperature advections having comparable magnitudes. Deformation fields associated with the growing disturbance produce temperature advection patterns that are highly favorable for concentrating the temperature gradient along the axis of the intensifying jet. The associated eddy heat fluxes are downgradient, with the net effect being to transport sensible heat westward (i.e., toward the eastern continent-western ocean side) as well as poleward across the central Pacific. In several aspects, the observed features resemble those required for baroclinic growth, although at the later stages, the spatial scale of the disturbance is larger and propagation speed smaller than for typical synoptic-scale developments.

On the basis of our analyses, we suggest that in many cases baroclinic processes are likely to play an important role in the developments, with barotropic processes (in particular, barotropic instability of the time-mean flow) more likely to provide significant contributions at a later stage of the development and in the maintenance of the anomalies once they are established. The observations therefore raise the distinct possibility that barotropic or equivalent barotropic models will be inadequate for simulating important aspects of the initial developments. We also speculate that, although variations in the response to tropical diabatic heating may play a role in enhancing the anomalous circulations, these variations may often be triggered by processes that are initiated at mid-latitudes.

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